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The climate response to CO_2 of the Hadley Centre coupled AOGCM with and without flux adjustment

by

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Abstract

Coupled ocean-atmosphere general circulation models have a tendency to drift away from a realistic climatology. The modelled climate response to an increase of CO_2 concentration may be incorrect if the simulation of the current climate has significant errors, so in many models, including ours, the drift is counteracted by applying prescribed fluxes of heat and fresh water at the ocean–atmosphere interface in addition to the calculated surface exchanges. Since it does not have a good physical basis, the use of this technique of "flux adjustment" itself introduces some uncertainty in the simulated response to increased CO_2. We find that the global-average temperature response of our model to CO_2 increasing at 1% per year is about 30% less without flux adjustment than with flux adjustment, while the geographical patterns are similar. The reduced response is due to more effective vertical transport of heat into the ocean, and a somewhat smaller climate sensitivity, both of which relate to differences in the initial climates. The response of the model with flux adjustment is more credible because it has a more realistic initial climate. The geographical similarity indicates that flux adjustment is not causing any gross distortion of the response.

Introduction

Coupled atmosphere-ocean general circulation models (AOGCMs) are the most comprehensive tools available for time-dependent climate-change experiments. Although the component atmosphere and ocean GCMs may separately produce acceptable simulations when forced at the surface with externally supplied realistic boundary conditions, significant errors in the climatology typically develop when the two are coupled to make an AOGCM. This presents a problem: if the simulation of present conditions (the "control" climate) is unrealistic, the response of the simulated system to a perturbation may not be correct, because the climate system is non-linear. A simple illustration of this can be found in the global-mean radiative equilibrium $S = \sigma T^4$, where S is the radiative flux into the climate system, and T the surface temperature required for equilibrium, assuming black-body emission. Increasing CO_2 increases S, and $\Delta T/\Delta S = T/4S$. Hence

if T is wrong in the control, ΔT will also be wrong. As we shall see, there are other non-linearities which are more important.

Many AOGCMs (see *e.g.* Gates *et al*, 1996) include "flux adjustments" (also called "flux corrections") (Manabe and Stouffer, 1988; Sausen *et al*, 1988), which are prescribed additional fluxes of heat, water and momentum applied at the air-sea interface in order to maintain realistic sea-surface conditions in the control. As well as giving greater confidence in the simulation of climate change, a better control climatology offers an additional advantage regarding model evaluation against observations. If there are large errors in the surface climatology, there will probably be consequential errors in many other simulated quantities (precipitation, cloud, sea-ice, etc.); hence comparison with observations will not be so useful in helping us to assess the performance of these other aspects of the model.

The flux adjustments depend only on location and time of year, not on the model state, so are exactly the same in the control and perturbed climates. Thus they do not explicitly introduce any additional feedbacks. However, they do give cause for concern because they may locally have values which are large compared with the physically computed atmosphere-ocean fluxes. If the flux adjustment, which is independent of climate, constitutes an important part of the surface budget, there is a distortion of the relationship between state variables and fluxes. Also, there is an implied transport between areas of negative and positive flux adjustment, again independent of climate. Both of these may affect the climate response.

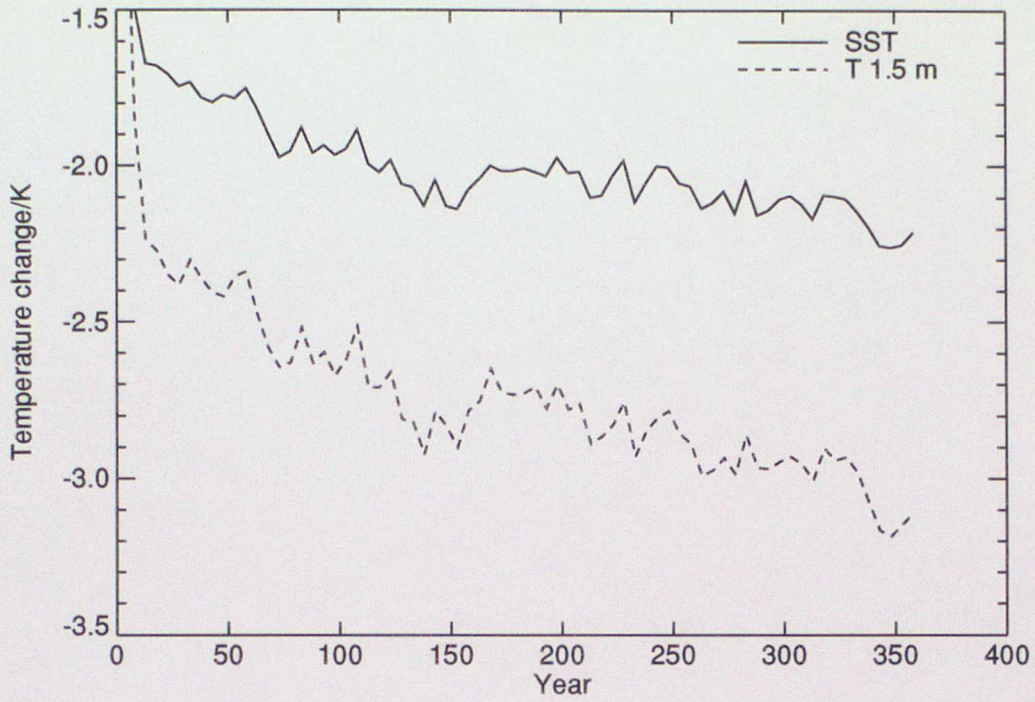
Some AOGCMs have been flux-adjusted in climate-change experiments (*e.g.* Cubasch *et al*, 1992; Manabe *et al*, 1992; Mitchell *et al*, 1995) and others not (*e.g.* Colman *et al*, 1995; Russell *et al*, 1995; Meehl and Washington, 1996). The climate response is qualitatively similar with and without, but there are differences in climate sensitivity and regional detail. It is not obvious whether any of these differences are due to flux adjustment, because the models differ in their formulations in many other respects which affect their climate response. To address this question, we have performed parallel climate-change experiments with and without flux adjustment using an AOGCM which is in all other respects exactly the same, allowing the first direct comparison of the response in the two cases. (A similar investigation has recently been carried out by Fanning and Weaver (1997) using an idealised AOGCM.) Here we report the basic results. Another paper will follow to describe the mechanisms in detail.

Model and experiments

We use a version of the Hadley Centre model known as HadCM2, which is normally run with heat and water flux adjustments to maintain realistic sea-surface temperatures and salinities, and has been used in various studies (Mitchell *et al*, 1995; Tett *et al*, 1996; Mitchell and Johns, 1997; Keen and Murphy, 1997). Johns *et al* (1997) describe the model, the spin-up, the derivation of the flux adjustments and the fluxadjusted control run, which we will refer to as "FC". FC has been run for over 1000 a and shows an upward drift in global-average surface air temperature of 0.16 mK a^{-1}. The global-average annual-average heat flux adjustment is 0.3 W m^{-2} (into the ocean), but locally it has a magnitude of up to 260 W m^{-2}.

An ensemble of four 80-year integrations were run with flux adjustment with CO_2 increasing at 1% per year compounded, reaching double its initial value after 70 years (Keen and Murphy, 1997). Radiative forcing increases linearly in time in these integra-

Figure 1: Timeseries of global-average temperature in the unfluxadjusted control run



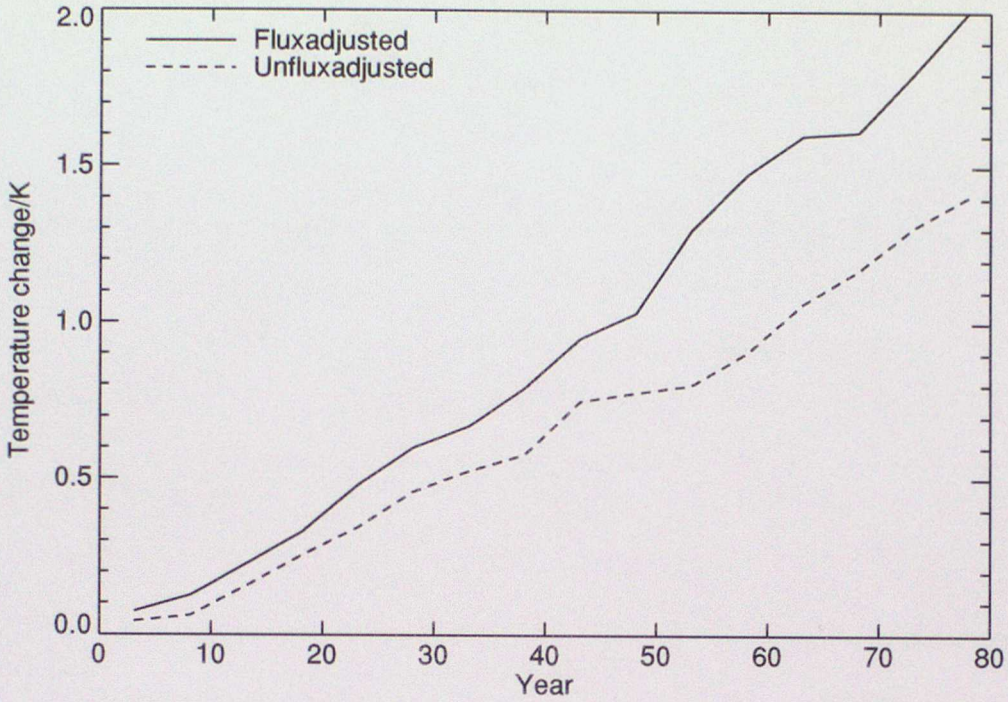
tions, since it depends logarithmically on CO_2 concentration. The integrations began at different arbitrary points of FC, so developed differently in detail. The ensemble-average is the best estimate of the climate change, and the intraensemble variability gives information about the statistical uncertainty of the result. We refer to this anomaly ensemble as "FA".

An integration ("UC") without flux adjustment was begun from an arbitrary point of FC, and an ensemble ("UA") of four integrations, exactly like those of FA but without flux adjustment, were started from various points of UC. The earliest began after 30 years, to allow for an initial adjustment, which we describe next. Because there is a substantial drift in UC, the signal of climate change without flux adjustment was obtained by subtracting corresponding years of UC from the UA integrations. For FC the drift is negligible so a long-term mean can be used to characterise its climate.

Unfluxadjusted control climate

In the first decade of UC, the climate cools rapidly (Figure 1). Over years 11–15 the global-average surface air temperature is 2.2 K lower than in FC, and the sea-surface temperature (SST) 1.7 K lower. The cooling is not the same everywhere. It is fairly uniform between 45°N and 45°S, where the amounts of low and medium-level cloud increase substantially and high cloud moves downwards. High northern latitudes cool more strongly, connected with an expansion of the area of sea-ice and snow and an increase in global-average surface albedo. There is a warming in the Southern Ocean. After the initial phase, the temperature in UC settles into a long-term slow downward drift, being 2.6 K cooler in surface air temperature than FC after 100 years; this continues during the 360 years for

Figure 2: Timeseries of global-average temperature change in response to increasing CO_2 with and without flux adjustment



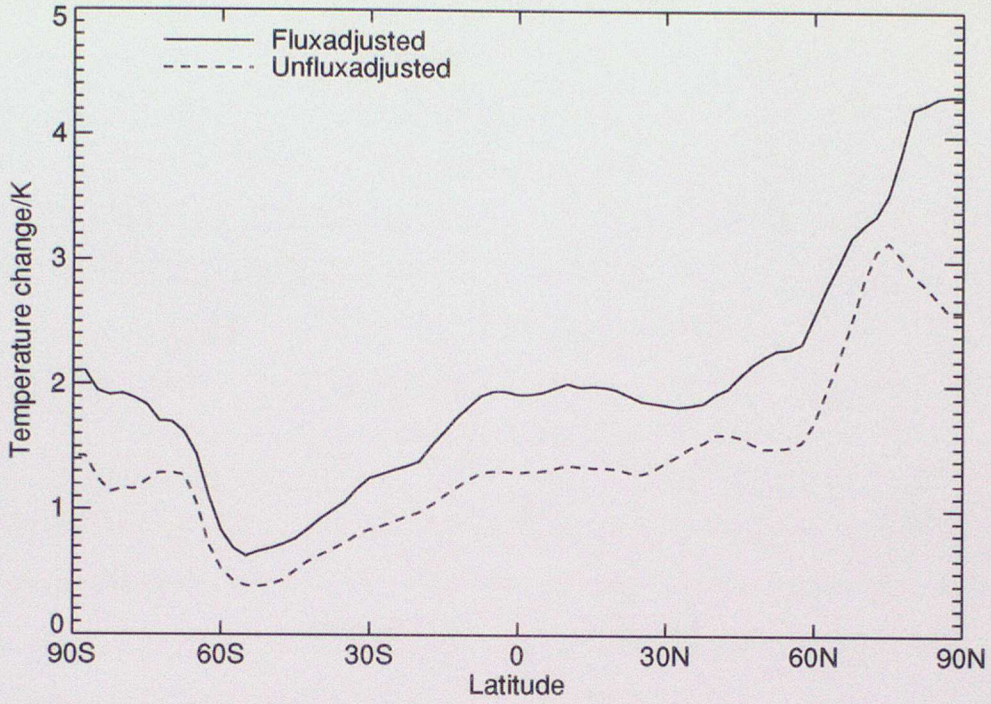
which UC has been run, and is fitted well by a linear trend of 1.9 mK a^{-1} for surface air temperature and 1.1 mK a^{-1} for SST. This drift is an order of magnitude larger than the warming tendency exhibited by FC.

It seems surprising that a global-average cooling as large as 2–3 K should result from a reduction in global-average heat input (the removed heat flux adjustment) of only 0.3 W m^{-2}, since a doubling of atmospheric CO_2, which produces a radiative forcing of ~ 3.5 W m^{-2} (after stratospheric adjustment), leads to a temperature rise of ~ 2.5 K (the IPCC "best estimate", Kattenberg *et al*, 1996). This comes about because the heat flux adjustment is *locally* large, so its removal causes large local temperature responses, but because there are geographically dependent feedbacks and long-range effects, there is not a uniform proportionality between local heat flux adjustments and temperature changes. Hence the positive and negative temperature changes do not balance to produce a small global average. The lack of proportionality can be seen in the small spatial correlations of the heat flux adjustment of about -0.2 with the surface air temperature change and -0.3 with the SST change. It appears that to eliminate the global-average temperature drift, we need the flux adjustments to be small *locally*, not just in the global average.

Climate change

The global-average temperature in both FA and UA rises fairly linearly with time, but less rapidly in UA (Figure 2). By the time when CO_2 has reached double its initial value (calculated as an average over years 61–80), the surface air temperature has risen by 1.76 ± 0.07 K in FA, but only 71% of this, 1.24 ± 0.06 K, in UA. (The uncertainty shown is the intraensemble standard deviation.) A comparison of zonal averages shows that the

Figure 3: Zonal-average global-average temperature change in response to increased CO_2 with and without flux adjustment



warming is reduced at all latitudes, by a fairly constant factor (Figure 3). The spatial correlation between the geographical distributions of surface air temperature change in FA and UA is 0.82, indicating the patterns are fairly similar. The correlation for SST is 0.69, which is smaller because part of the common signal in the surface air temperature is the land-sea contrast in the warming (greater over land). There are local differences, especially in northern high latitudes, which can partly be related to specific local differences in the climatologies of FC and UC.

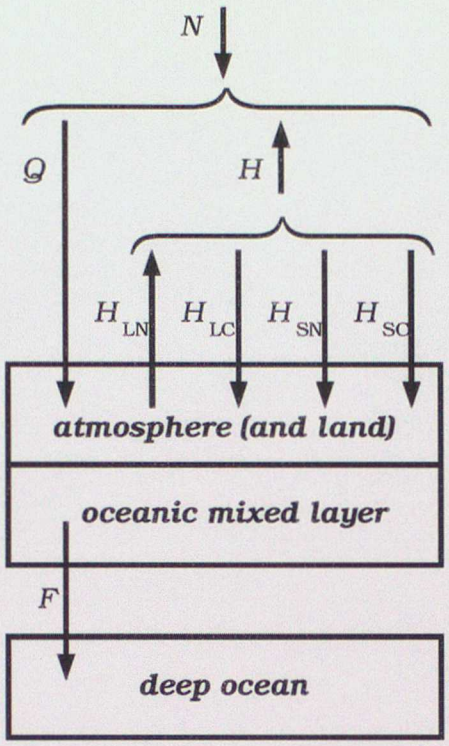
When considering timescales of several years, the heat capacity of the atmosphere and mixed layer can be neglected. Therefore if Q is the imposed radiative forcing of the climate system,

$$Q = H + F \tag{1}$$

(see Figure 4). H is the increase in radiation to space at the top of the atmosphere (TOA), both longwave and shortwave, including feedbacks, resulting from the warming. F is the heat flux from the mixed layer into the deep ocean. When the climate regains a steady state, there must be no flux on average from the mixed layer to the deep ocean *i.e.* $F = 0 \Rightarrow Q = H$. Thus, H characterises the steady-state response of the climate system to Q, and F the transient behaviour. It may not be possible to attain a steady state in practice in an AOGCM, because the timescale on which the deep ocean equilibrates is millenia, and computing resources may not be available to perform such an integration.

Greenhouse gases inhibit the emission of longwave radiation to space, so the forcing Q they cause appears as a reduction in outgoing longwave radiation (OLR) at the TOA. Since H is an increase in TOA outgoing radiation, the change in the net downward flux at the TOA is $N = Q - H$. Assuming that H depends linearly on the temperature change ΔT of the climate system (in practice the global-average surface temperature), we write

Figure 4: Relationships between global-average fluxes



$H = \lambda\Delta T$, with λ a constant "climate response factor" (see *e.g.* Hansen *et al*, 1984; Dickinson, 1986). Hence $\lambda\Delta T = Q - N$. This enables us to obtain λ from N, Q and ΔT at any time; for the time of doubled CO_2 we find $\lambda = 1.32 \pm 0.06$ W m^{-2} K^{-1} for FA, 1.79 ± 0.10 W m^{-2} K^{-1} for UA (Table 1).

The conventional "climate sensitivity" $\Delta T_{2\times}$ is the global-average temperature difference between the control and a steady-state climate with a doubled concentration of CO_2. For a steady state $Q = H$, so if the forcing is $Q_{2\times}$, knowing λ we can obtain $\Delta T_{2\times}$ as $Q_{2\times}/\lambda$, without having integrated the model to a steady state. For FA, $\Delta T_{2\times} = 2.64 \pm 0.12$ K; for UA, 1.94 ± 0.19 K (Table 1). Both are within the generally accepted range (Kattenberg *et al*, 1996).

H is a combination of (Figure 4):

H_{LN} Change in clear-sky OLR (*i.e.* computed with no cloud) arising from the basic radiative response of the system to the higher temperature and from the water-vapour feedback.

H_{LC} Change in longwave cloud radiative forcing (CRF). Longwave CRF is the difference between clear-sky OLR and OLR with normal cloud, and is positive in the global average because clouds inhibit longwave radiation to space.

H_{SN} Change in clear-sky net downward shortwave (*i.e.* a reduction in upward shortwave), resulting from reduced surface albedo.

H_{SC} Change in shortwave cloud radiative forcing. Shortwave CRF is defined like longwave CRF, and is negative because clouds reflect sunlight.

Table 1: Comparison of climate response

	FA	UA
Change in global-average temperature ΔT/K	1.76	1.24
Radiative forcing $Q = Q_{2\times}$/W m^{-2}	3.47	3.47
Change in net radiation at TOA N/W m^{-2}	1.16	1.25
$H = Q - N = \lambda \Delta T$/W m^{-2}	2.31	2.22
λ/W m^{-2} K^{-1}	1.32	1.79
$\Delta T_{2\times}$/K	2.64	1.94
λ_{LN}/W m^{-2} K^{-1}	1.82	1.81
λ_{LC}/W m^{-2} K^{-1}	0.03	−0.28
λ_{SN}/W m^{-2} K^{-1}	0.71	0.85
λ_{SC}/W m^{-2} K^{-1}	−0.24	−0.55
$F = N = \kappa \Delta T$/W m^{-2}	1.16	1.25
κ/W m^{-2} K^{-1}	0.66	1.01

The first of these is defined upward, the others downward, so $H = H_{\mathrm{LN}} - H_{\mathrm{LC}} - H_{\mathrm{SN}} - H_{\mathrm{SC}}$. Each of them is assumed separately proportional to ΔT, so the combined $\lambda = \lambda_{\mathrm{LN}} - \lambda_{\mathrm{LC}} - \lambda_{\mathrm{SN}} - \lambda_{\mathrm{SC}}$. When $\lambda_{\mathrm{LC}}, \lambda_{\mathrm{SN}}, \lambda_{\mathrm{SC}}$ are positive, they give downward fluxes, which reinforce the radiative forcing, and constitute a positive feedback on the warming.

FA and UA have similar clear-sky longwave feedback λ_{LN} (Table 1), which means that the temperature difference between FC and UC is not large enough to have an important influence on the σT^4 term or the water-vapour feedback. The positive clear-sky shortwave feedback λ_{SN} is slightly stronger in UA than in FA, presumably because of the greater initial area of sea-ice and snow. The main difference in the response derives from the cloud feedbacks; the explanation must lie in the different cloud climatologies of FC and UC. λ_{SC} and λ_{LC} both become more negative, the most important influence being reduced positive feedbacks at low latitudes. For the shortwave, this could be because UC has much more low and medium-level cloud than FC. Consequently, the reduction of high cloud resulting from the warming has less effect on the planetary albedo. For the longwave, a possible explanation is that the increase in the upper-level cloud water path is less in UA because the temperature is lower.

Assuming the heat capacity of the mixed layer and atmosphere is small, the net downward flux N at the TOA is nearly equal to the net heat flux F into the deep ocean (Figure 4). Timeseries of N show that it grows roughly linearly in time (Figure 5). Since ΔT also increases approximately linearly in time, there is a proportionality $F = \kappa \Delta T$. We do not expect this *a priori*, and it does not have the general applicability of $H = \lambda \Delta T$. It obviously fails as the system approaches a steady state, when $F \to 0$ but ΔT tends to a non-zero final value. Nonetheless, it is useful in the present analysis.

It is surprising that F is very similar in FA and UA; again, there is no reason to expect this, which means that the heat taken up by the deep ocean is the about same in the two experiments, despite the fact that the surface warming is 30% less in UA. In fact, the increase in heat content of the entire ocean averaged over years 61–80 in UA is only 8% less than in FA. A calculation of the distribution of this heat with depth shows that, on average, it penetrates more deeply in UA, meaning that the vertical mixing processes are more effective in the unfluxadjusted ocean (Figure 6). The mechanisms for this will be discussed in a later paper.

Figure 5: Timeseries of global-average net downward radiation at the top of the atmosphere

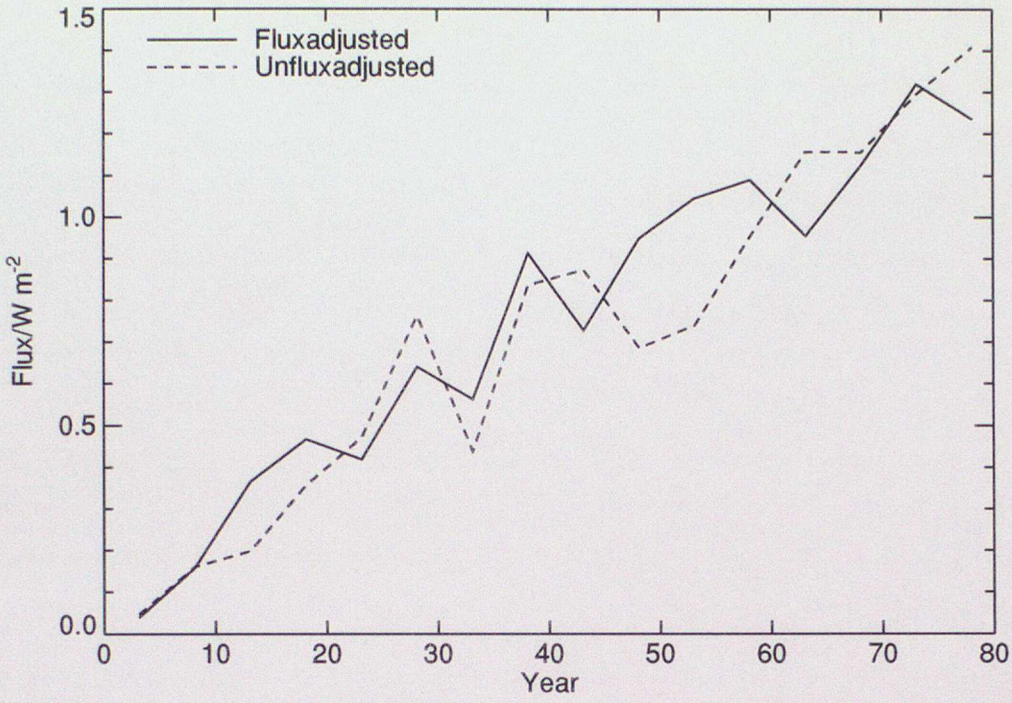
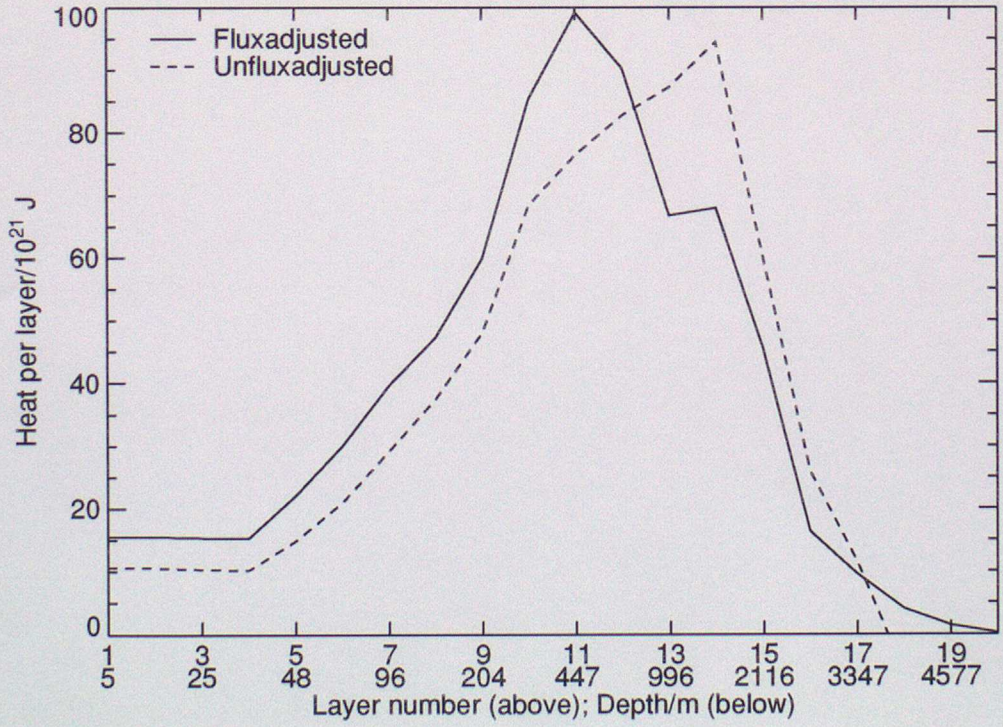


Figure 6: Storage of additional heat in the ocean as a result of increased CO_2 with and without flux adjustment



The constant κ quantifies the transfer of heat from the surface to the deep ocean. Since F is the same but ΔT is less in UA, κ must be greater, consistent with the conclusion drawn from the vertical distribution of heat uptake. Calculating $\kappa = F/\Delta T$ yields 0.66 ± 0.04 W m^{-2} K^{-1} in FA, 1.01 ± 0.16 W m^{-2} K^{-1} in UA.

In terms of ΔT, Equation (1) becomes

$$Q = (\lambda + \kappa)\Delta T. \tag{2}$$

This shows that the surface temperature is determined by the effectiveness of heat loss to space and into the deep ocean. Because λ and κ are bigger in the unfluxadjusted model, ΔT is smaller. The differences between the two models in λ and κ are similar, so the smaller climate sensitivity and the increased vertical transport are of comparable importance in reducing ΔT.

Conclusion

Although the global-average flux adjustment is small, substantial local and global-average climate drift occurs in HadCM2 without it. The drift is amplified by local feedbacks, suggesting that the flux adjustment has to be small locally, as well as on average, to eliminate global-average drift. From the comparison of response experiments, we find that global-average temperature change resulting from increasing CO_2 is less in our model without flux adjustment than with, everything else in the model being the same. The geographical patterns of temperature change are generally similar. The altered response is not, however, due directly to the flux adjustment, but to the considerably different initial states of the atmosphere and ocean; it comes about mainly through a reduced climate sensitivity caused by differences in the cloud climatology, and increased effectiveness of vertical heat transport in the ocean. Insofar as the fluxadjusted initial state is closer to observations, this comparison provides support for the use of the technique of flux adjustment; we have more confidence in the climate response from the fluxadjusted model because its control state is more realistic. At this stage in the analysis we cannot rule out the possibility that the flux adjustment does also directly influence the climate response, although apparently it does not introduce any gross distortion. Our best strategy remains to try to improve the model to reduce the flux adjustments, repeating the comparison of climate response with and without the technique until they converge.

Acknowledgements

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